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The volcanotectonic evolution of Flores Island, Azores (Portugal)

J.M.M. Azevedo *, M.R. Portugal Ferreira

Dep. de Ciências da Terra, Universidade de Coimbra, 3000-272 Coimbra, Portugal

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Abstract

From the study and interpretation of the volcanic products and structures of Flores Island, we infer that its volcanic history was dominated by two major periods: (1) proto-insular volcanism, which includes all the submarine and emergent activities; and (2) insular volcanism, consisting exclusively of subaerial eruptions. The first period includes two phases: (1) the oldest (2.2 to 1.5 Ma) of shallow submarine volcanism; (2) the youngest (1.0 to 0.75 Ma) includes emergent volcanism. Throughout the second period, three volcanic stages are recognized: (1) the first one (0.7 to 0.5 Ma) includes the most voluminous volcanism, balanced between effusive and explosive events; (2) an intermediate stage (0.4 to 0.2 Ma) that involves a larger number of small-scale feeder centres, with effusive eruptions prevailing; (3) the third stage is the latest volcanic activity of the Island (≈0.003 Ma), with strombolian and subsequent phreatic and phreatomagmatic activity centred at four volcanic vents.

From 1.0 Ma to the present, the volcanotectonic development of Flores Island also reflects the operation of two major tectonic processes: (1) a marked volcanotectonic uplift during the first stage (1.0 to 0.55 Ma); and (2) subsidence, particularly intensive in the island's central area, which led to the subsequent formation of two and perhaps three large *calderas* during the period 0.55 to 0.4 Ma.

Keywords: volcanic island; volcanic complex and unit; volcanostratigraphy; subsidence and uplift

1. Introduction

Before the mid-1980s, no systematic and comprehensive geological study had been carried out on Flores Island. Our programme started in 1985 with the purpose of improving the 1:25,000 geological map of the island by Zbyszewsky et al. (1968). Initial studies focused on the older volcanic formations (Azevedo et al., 1986; Azevedo, 1988, 1990; Azevedo et al., 1991). Later mapping was extended to the entire island and island-wide volcanostratigraphy was defined. A new 1:15,000 volcanological map includes considerable new geochronologic, petrographic and geochemical data, as well as description of the tectonic and geomorphological histories (Azevedo, 1999, vol. 1; Azevedo and Ferreira, 1999). The volcanostratigraphy and related structural interpretation supported the further studies focused on hydrogeological modelling (Azevedo, 1999, vol. 2).

This paper provides a refined interpretation of the volcanic growth and the structural evolution of Flores, within the framework of the entire Azores region.

2. Geographic, geotectonic and geochronologic settings

Flores and Corvo islands constitute the western group of the Azores Archipelago (Fig. 1), which is near the middle of Atlantic Ocean and forms a 600 km-long

* Corresponding author. Tel.: +351 239 860 500; fax: +351 239 860 501.
E-mail address: jazevedo@dct.uc.pt (J.M.M. Azevedo).

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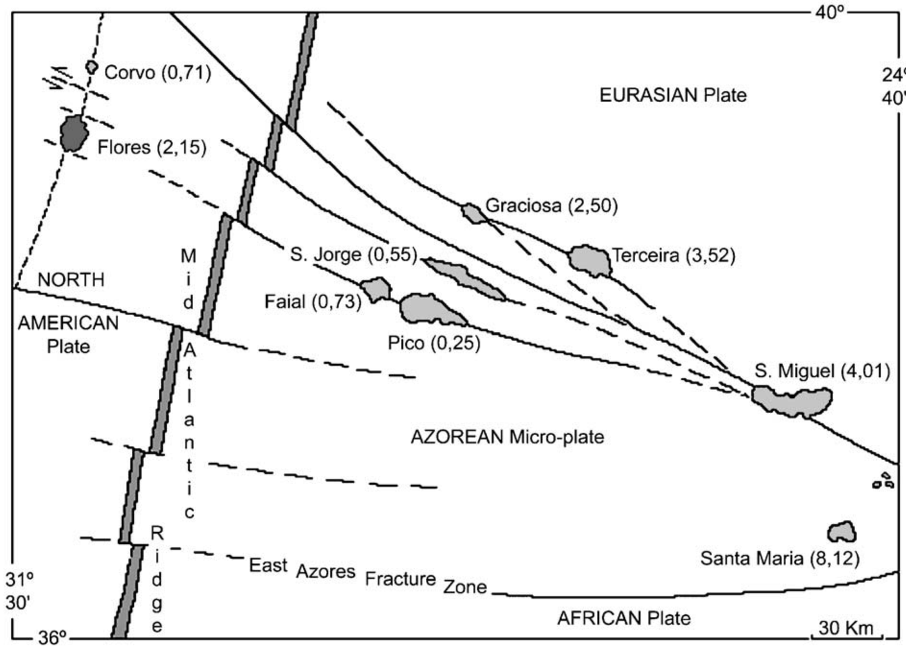

Fig. 1. Geographic and geotectonic setting of Azores Archipelago (adapted from Forjaz, 1988; Baptista et al., 1999) with the oldest radiometric ages (in parentheses; Ma) for each island (geochronological data from Abdel-Monem et al., 1968, 1975; White et al., 1976; Ferraud et al., 1980; Ferreira and Martins, 1983; Feraud et al., 1984; Forjaz, 1988; Azevedo et al., 1991; Azevedo, 1999; Nunes, 1999; Azevedo et al., 2003).

belt trending in a WNW direction. All nine islands (and related seamounts) of this archipelago are within a zone where three lithospheric plates (North American, Eurasian, and African) meet. Most of the islands are located along the fracture-zone extensions of transform faults of the Mid-Atlantic Ridge (MAR). Unlike the other seven islands, Flores and Corvo are on the North American Plate (Fig. 1). The islands of Central Group (Faial, Pico, S. Jorge, Terceira and Graciosa) and Eastern Group (S. Miguel and Santa Maria) are within a transition zone (called the Azorean micro-plate, after Forjaz, 1988) between the Eurasian and African Plates.

Several interpretations and models for the tectonic setting and geodynamic regime of the Azorean region have been proposed mainly on the basis of neotectonic, seismotectonic, GPS data and paleomagnetic interpretations (see for example: Krause and Watkins, 1970; McKenzie, 1972; Machado et al., 1972; Searle, 1980; Hirn et al., 1980; Machado et al., 1983; Buforn et al., 1987; Madeira and Ribeiro, 1990, 1992; Freire Luís et al., 1994; Madeira, 1998; Baptista et al., 1999).

Existing geodynamic models are mainly or exclusively focused on the islands of the Central Group (i.e., Azores microplate) and generally lack a detailed discussion of the tectonic regime for the Western Group islands of Flores and Corvo. This lack reflects the situation that these two islands: (1) are west of the MAR; (2) are aligned north–south, which is parallel to the MAR and nearly perpendicular to the alignment of the other Azorean islands; (3) have no record of significant historical seismicity; and (4) are subjects of few geophysical surveys.

Nonetheless, the results of several studies (Serughetti and Rocha, 1968; Krause and Watkins, 1970; Blakely, 1974; Azevedo, 1988, 1990; Azevedo et al., 1991; Bastos et al., 1993; Freire Luís et al., 1994; Azevedo and Ferreira, 1995; Baptista et al., 1999; Azevedo, 1999; Azevedo and Ferreira, 1999) support the following observation about Flores and Corvo volcanotectonic setting:

– Both islands are high subaerial parts of a single large mostly submarine edifice, built on a 9.0 to 10.0 Ma oceanic crust (Blakely, 1974; Needham and Francheteau, 1974; Freire Luís et al., 1994).
– The tectonic setting and volcanic construction of both islands are likely related to the geodynamics of the MAR and associated transform faults. However, this inferred that structural control is clearly less evident than those for the central and eastern Azorean islands.
– A westward displacement of Corvo from Flores at about 1 cm/year (Baptista et al., 1999), together with the near-linear E–W north and south coastlines of

Flores, reflect structural control by MAR transform faults, both in the past and at present.

- The main tectonic lineaments on Flores island are sinistral strike-slip N30–40°W and normal N20–30°E faults (Azevedo, 1999) (Fig. 3). The N–S structural alignment emphasized in Forjaz (1988) has a clear secondary expression on the island subaerial domains.
- Development of important vertical tectonic movement affecting the crustal region of Flores Island since 1 Ma ago. As a matter of fact, it should be noted that a submerged island (in present-day it corresponds to a seamount at the depth of 450 m) located 50 km westwards of Flores Island shows a tectonic evolution (Ryall et al., 1983) different than that for Flores (see Section 5); while in Flores uplift dominated, the seamount subsided at about the same rate, supporting the concept of an isostatic compensation between the two adjacent blocks of the oceanic crust.

The geochronological data (Table 1) confers to the Flores oldest subaerial lavas an age of about 2.2 Ma (Azevedo, 1988; Azevedo et al., 1991), which means

Table 1
Radiometric ages and chronostratigraphy of volcanic rocks from Flores Island, Azores

Sample	Location on Fig. 3	Lithology-Structure	K (%)	Ar^{40} rad. (^{cc}STP/G.10^{-8})	Ar^{40} atm. (%)	Age (M.a.)	Volcanic unit
1 [a]	1	Carbonized wood				0.0029±0.0001	UC3
2 [a]	2	Carbonized wood				0.0030±0.0001	
7–95 [b]	3	Basalt–Lava-flow	1.28	1.089	96.84	0.23±0.12	UC2
6–95 [b]	4	Basalt–Lava-flow	1.19	1.250	91.37	0.27±0.11	
31–88 [b]	5	Basalt–Lava-flow	1.31	2.465	91.98	0.35±0.08	
2–95 [b]	6	Basalt–Lava-flow	1.44	2.899	80.23	0.40±0.08	
FL29 [c]	7	Mugearite–Lava-flow	2.65	2.150	89.00	0.55±0.12	UC1
246–87 [b]	8	Basalt–Lava-flow	1.06	2.323	94.91	0.56±0.12	
75–81 [b]	9	Mugearite–Lava-flow	3.24	6.968	82.07	0.57±0.13	
4–95 [b]	10	Hawaiite–Lava-flow	2.74	6.993	88.65	0.57±0.09	
123–85 [b]	11	Trachyte–Dike	4.35	8.763	92.55	0.57±0.05	
178–86 [b]	12	Basalt–Lava-flow	1.09	0.175	98.75	0.58±0.16	
18–81 [b]	13	Trachyte–Dike	4.33	9.851	71.86	0.59±0.05	
179–86 [b]	14	Basalt–Lava-flow	1.65	3.788	93.23	0.59±0.05	
8–95 [b]	15	Hawaiite–Lava-flow	2.35	5.405	65.12	0.59±0.08	
FL8 [c]	16	Mugearite–Lava-flow	1.93	1.700	79.00	0.61±0.09	
150–86 [b]	17	Basalt–Lava-flow	0.72	1.838	90.38	0.62±0.48	
FL22 [c]	18	Trachyte–Lava-flow	4.47	4.090	60.00	0.62±0.05	
118–85 [b]	19	Hawaiite–Dike	2.16	5.877	89.57	0.64±0.05	
153–86 [b]	20	Basalt–Lava-flow	0.88	2.198	92.94	0.64±0.28	
183–86 [b]	21	Basalt–Lava-flow	1.23	0.186	99.66	0.64±0.24	
140–85 [b]	22	Hawaiite–Dike	2.66	6.812	86.27	0.66±0.03	
245–87 [b]	23	Basalt–Lava-flow	1.59	3.507	91.08	0.74±0.15	BC1
21–88 [b]	24	Hawaiite–Lava-flow	2.39	9.390	81.89	0.86±0.04	
134–85 [b]	25	Basalt–Lava-flow	0.92	1.349	97.99	0.96±0.19	
146–85 [b]	26	Basalt–Lava-flow	1.58	7.573	95.85	0.97±0.32	
22–81 [b]	27	Mugearite–Dike	3.09	12.012	71.54	0.98±0.04	
176–86 [b]	28	Basalt–Lava-flow	1.70	4.410	87.82	1.01±0.08	
167–86 [b]	29	Basalt/Hawaiite–Lava-flow	1.84	6.289	91.98	1.53±0.07	BC2
16–81 [b]	30	B-Neck	1.53	2.590	98.16	1.57±0.45	
201–86 [b]	31	Basalt/Hawaiite–Lava-flow	1.81	11.866	73.34	1.79±0.05	
202–86 [b]	32	Hawaiite–Lava-flow	2.11	16.611	81.26	1.88±0.67	
226–87 [b]	33	Basalt/Hawaiite–Lava-flow	1.85	12.890	77.70	1.96±0.06	
128–85 [b]	34	Basalt–Lava-flow	0.44	−0.708	99.96	2.04±0.79	
233–87 [b]	35	Hawaiite–Lava-flow	2.00	15.131	90.56	2.11±0.62	
93–81 [b]	36	Basalt–Lava-flow	1.65	15.423	71.73	2.16±0.16	

[a] C^{14} determinations from Morisseau (1985).
[b] K/Ar ages from Azevedo et al. (1986, 1991) and Azevedo (1990, 1999); the determinations were obtained in the Geochronological Laboratory, Coimbra University.
[c] K/Ar ages from Ferraud et al. (1980).

that only Santa Maria, S. Miguel, Terceira and Graciosa Islands have older subaerial formations (Fig. 1).

3. Volcanic rocks and volcanostratigraphy

As with the other Azorean Islands, the Flores volcanic activity has included different processes and eruptive styles and dominantly involved basaltic and hawaiitic lavas. However, differentiated products, such as mugearites, benmoreites and trachytes, are also reported (Zbyszewsky et al., 1968; Torre de Assunção et al., 1974; Morrisseau and Traineau, 1985; Azevedo, 1999).

Present-day outcrops on the island include products and structures of the subaerial volcanism as well as of the emergent and submarine eruptions. Accordingly, the volcanic lavas and deposits were grouped into two major Complexes (Azevedo et al., 1986):

(1) The Base volcanic Complex (BC), which includes all the products and structures that resulted from the submarine and emergent volcanism.
(2) The Upper volcanic Complex (UC), which includes all the volcanic products of the subaerial volcanic activity.

Geologic mapping at a scale of 1:15,000 (Azevedo, 1999) – supported by radiometric ages (K/Ar data from Ferraud et al., 1980; Azevedo et al., 1986; Azevedo, 1990; Azevedo et al., 1991; Azevedo, 1999; ^{14}C data from Morrisseau and Traineau, 1985; Table 1) and by petrographic and geochemical data (Torre de Assunção et al., 1974; Azevedo, 1999, 2003) – allows us to subdivide both Complexes and define a relatively detailed volcanostratigraphy (Table 2).

The BC formations crop out along the lower levels of coastal or paleo-coastal cliffs (Fig. 3) and are subdivided into Base Complex 1 (BC1) and Base Complex 2 (BC2), of Plio-Pleistocene age. The BC rocks are mainly volcaniclastic deposits (a mixture of hyaloclastite and hydroclastite, as defined by Batiza and Wite, 2000) of breccias and tuffs; these show intensive and pervasive palagonitization, and a great variability in texture, dimension (0.005 to 1.5 m), and shape of the clasts. Pyroclasts are more common in BC1 deposits and autoclasts prevail within BC2 formations. Massive structures and intensive lithification are very pervasive in the breccias. An incipient and sub-horizontal lamination in the tuff and stratification in the coarser deposits are common. Some basaltic or hawaiitic lava-flows (Fig. 2) are interbedded in the middle and upper breccias and tuffs (Azevedo, 1988).

The UC is subdivided into Upper Complex 1 (UC1), Upper Complex 2 (UC2) and Upper Complex 3 (UC3). The upper Unit (UC1) was formed between 0.66 and 0.55 Ma (Table 1) and consists of extensive and sometimes very thick lava flows alternating with subordinate pyroclastic deposits. The UC1 rocks evolved from basaltic to trachytic compositions (Fig. 2). The middle Unit (UC2), whose age ranges from 0.4 to 0.2 Ma, is composed of basaltic and hawaiitic lava flows (Fig. 2) and associated pyroclastic deposits. The

Table 2
Volcanostratigraphy of Flores Island (adapted from Azevedo, 1999)

Group (complex)	Volcanic unit	Volcanic sub-unit	Geochronology (Ma)
Upper complex—UC	Upper—UC3	Phreatomagmatic deposits—H	0.002
		Strombolian scoria cones—G	0.003
	Middle—UC2	Intermediate hawaiites—F	0.22
		Intermediate basalts—E	0.40
	Lower—UC1	Inferior benmoreites-trachytes—D	0.55
		Inferior hawaiites-mugearites—C	
		Inferior basalts-hawaiites—B	
		Inferior basalts—A	0.67
Base complex—BC	Upper—BC1	Palagonitized breccias and tuffs—a	0.70–0.80
		Palagonitized basalts and hawaiites—b	1.0–1.5
		Palagonitized breccias—c	≅1.8
	Lower—BC2	Palagonitized basalts—d	2.0–2.2
		Palagonitized volcaniclastic deposits—e	

-------- Passage without volcanic quiescence and depositional unconformity.
– – – – – Passage with volcanic quiescence and/or depositional unconformity.

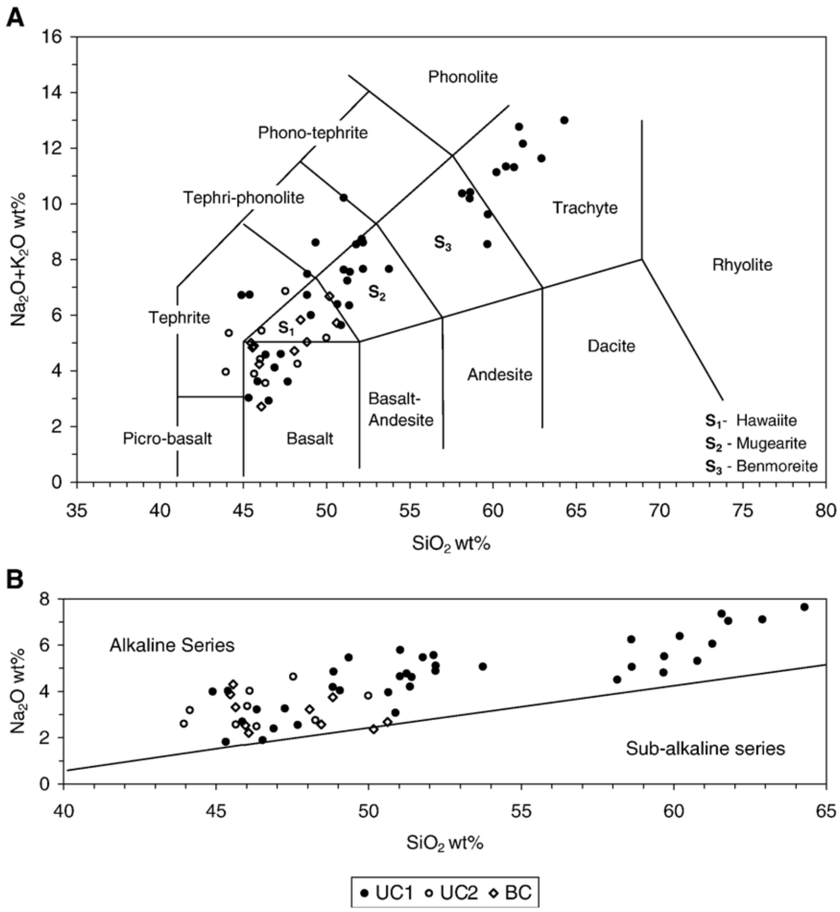

Fig. 2. Geochemical characterization of the lavas from Flores Island (geochemical data from Torre de Assunção et al., 1974; Azevedo, 1999); BC — Base volcanic Complex, UC — Upper volcanic Complex): (A) Total Alkali–Silica (TAS) classification (after Le Bas et al., 1986); (B) Silica–Na_2O variation diagram (after Middlemost, 1975).

upper Unit (UC3) was emplaced between 0.004 and 0.003 Ma (Morrisseau and Traineau, 1985) and locally includes strombolian pyroclastic cones and a widespread ash mantle enriched with lithic clasts.

Periods of volcanic quiescence (or volcanic gaps) within the overall volcanic evolution of Flores Island (Table 2 and Fig. 5) are confirmed by field observations — not only by the absence of volcanic rocks, which in turn results in gaps in geochronological data, but also by the occurrence of pervasive paleosoils and/or depositional, lithological and structural unconformities within the volcanic sequences.

Taking into account the volcanostratigraphy, the petrochemical data from the effusive volcanic rocks of Flores Island demonstrate: (1) a great predominance of alkaline basalt and hawaiitic compositions; (2) a continuous petrological and geochemical sequence ranging from basalt to trachyte (Fig. 2A) recording a volcanogenetic differentiation process, which happened between about 0.75 and 0.40 Ma; (3) the subsequent volcanism (UC2 and UC3 eruptive episodes) involved exclusively basaltic and hawaiitic compositions; and (4) a pervasive sodic alkaline tendency (Fig. 2B).

4. Volcanic setting and build-up processes

The following reconstruction of volcanic history is mainly based on the interpretation of the primary lithological and structural *facies* of the eruptive products. However, the study of the secondary (or alteration) *facies* was particularly important for the interpretation of the BC volcanic setting.

4.1. Base complex

The diversity of products and structures present in the BC volcanic rocks points to a wide range of extrusive styles and depositional processes. Taking into account the scarcity of BC outcrops (Fig. 3) and the intensive secondary alteration (particularly of the volcaniclastic breccias and tuffs), the deciphering of their extrusional and depositional mechanisms is difficult.

The widely scattered occurrences of the BC outcrops (Fig. 3) argue for the existence of multiple feeder centres. The typically short length and thickness of most of the BC volcanic deposits suggest the development of small-to-medium scale volcanic centres. The coarse symmetrical distribution of the BC formations along a NNW–SSE axis suggests that structural lineaments with a similar trend possibly may have played an important role in the localization and evolution of the BC volcanic centres.

The occurrence of pillow lavas and the intensive palagonitization of much of the BC volcaniclastic deposits (hydroclastic deposits) indicate a regime of submarine and emergent volcanism. This corresponds to the proto-insular volcanism or the intermediate water/shoaling stage, as defined by Schmidt and Schmincke (2000).

The very high percentage of volcaniclastic deposits within the BC Units indicates the involvement of highly efficient fragmentation processes. Moreover, the great variability in texture, dimension, and shape of the clasts supports the occurrence of both pyroclastic and autoclastic fragmentation, implying different degrees of lava–seawater interaction and consequently the occurrence of several eruptive styles. The predominance of pyroclastic deposits over autoclastic deposits and lava flows indicates a prevalence of explosive volcanism. Only two predominantly effusive periods are recorded, one by Sub-unit b of BC1 (Palagonitized Basalts and hawaiites) and another by Sub-unit d of BC2 (Palagonitized Basalts) (Table 2).

The morphometric and textural characteristics of the BC pyroclasts reflect explosive fragmentation from both magmatic and phreatomagmatic processes. The common presence of autoclasts and hyaloclastites within the BC breccias suggests an effective and widespread non-explosive fragmentation by autobrecciation (or flow fragmentation, as defined by Cas and Wright, 1987) and thermal granulation or quenching processes.

The stratigraphic sequence of the BC breccias and tuffs marks progression from submarine to emergent volcanic style, i.e., the transition from the intermediate to shoaling stage as defined by Schmidt and Schmincke (2000). Three stages of emergence are apparent: (1) the stage of dominantly nonexplosive fragmentation, reflecting deep submarine volcanism; this activity is recorded by the Sub-unit e of BC2 (Palagonitized volcaniclastic deposits, Table 2); (2) the stage of dominant hydroclastic explosive activity; this volcanism occurs below the volatile fragmentation depth (VFD, after Fisher and Schminke, 1984); and (3) the stage of the “sea-level” volcanism (or emergent volcanism) alternating between dry and temporary flooding of the vent (as defined by Kokelar, 1986; Cas and Wright, 1987); it corresponds to the most explosive BC volcanism, and it is recorded by the upper BC Sub-unit (Palagonitized breccias and tuffs, Table 2).

Moreover, lateral and vertical changes within the internal structure of the BC breccias and tuffs indicate several transportation-depositional mechanisms occurred within various depositional settings on the ocean floor. The considerable thickness (in some cases, about 100 m) and the high incidence of massive internal structure for the majority of the BC deposits point to a quick succession of submarine volcaniclastic flows. The large amount of clasts supported by a palagonitized matrix suggests the development of submarine *lahars* and debris-flow processes (Azevedo, 1990; Azevedo et al., 1991). On the other hand, the volcaniclastic deposits without matrix-support reflect the progression of coarse grain-flows and collapses. The volcanic tuffs, and particularly the usual occurrence of cross bedding and double-grading structure, seem to be mainly the result of mudflows and turbidity currents (Azevedo, 1988).

The scarcity of pillow structures in the BC formations is a consequence of the low productivity of the effusive eruption, which is evidenced by the small thickness of the lava flows, and probably as well as the steep slope of the ocean floor.

4.2. Upper complex

The emplacement of Upper Complex (UC) began about 0.7 Ma ago within a very restricted area (approximately inside the present-day island area) and it comprehended an eruptive process mainly based on a vertical overlapping of the volcanic events and products. The descriptions of the three UC Units volcanism follow.

4.2.1. Upper complex 1 (UC1)

The growth of this UC Unit generally involved volcanic processes with remarkable spatial and temporal continuity. However, there was a gradual evolution both in the volcanic activity and on the number and dimension of the feeder centres. Therefore, the deposition of UC1 may be divided into two sequential phases.

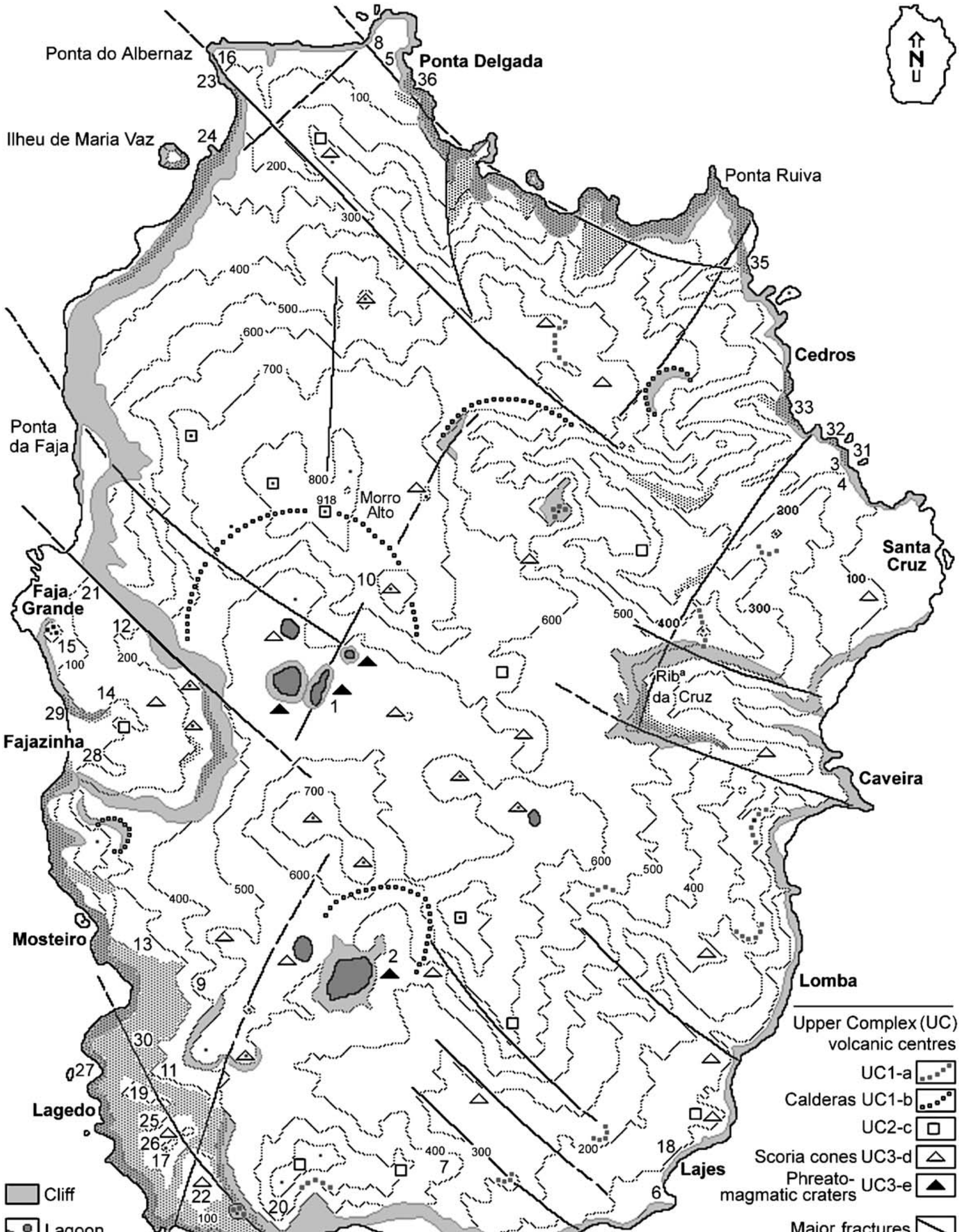

Fig. 3. Sketch map of Flores Island, showing: (1) Upper Complex major feeder centres (a, UC1 centres; b, UC1 affiliated *calderas*; c, UC2 centres; d, UC3 scoria cones; e, UC3 phreatomagmatic craters), (2) major fractures and (3) Base Complex outcrops. Numbers indicate dated lavas (see Table 1).

Phase 1

Feeder centres	•Number and type: Two or three craters of large diameter (Figs. 3 and 4) •Morphology: High lateral extension; low to medium elevation •Localization: In the centre and on the NE and S peripheries of the island
Volcanic system	•Polygenetic: Association of small-to-medium-scale central volcanoes (as defined by Walker, 2000)
Volcanic activity	•Explosivity: Gradual evolution from about 50% to less than 30% •Eruptive style: Changeable (but mostly hawaiian–strombolian)
Structural setting	•Fracture systems: The volcanic centres were rooted on the intersection of N30°–40°W with N20°–30°E fractures systems; most fractures were filled with volcanic dikes •Regional geodynamic regime: Distensive (prolongation from the previous tendency)
Rock record	•Sub-units A and B from UC1 (Table 2)
Time interval	•0.7 to about 0.6 Ma (Table 1)

Phase 2

Feeder centres	•Number and dimension: Association of seven or eight new small dimension craters to the previous large centres (Fig. 3); intrusion of several volcanic necks •Localization: In the centre and along the periphery of island
Volcanic system	•Polygenetic: Association of small-scale central-volcanoes with asymmetric cones and volcanic necks
Volcanic activity	•Explosivity: Gradual evolution from 30% to about 5% •Eruptive style: Progressive evolution from hawaiian–strombolian to plinian activity
Structural setting	•Fracture systems: The same as for phase 1 (fractures N30°–40°W with N20°–30°E) in association with small-scale fractures related to the large volcanic centres
Rock record	•Sub-units C and D from UC1 (Table 2)
Time interval	•0.6 to 0.55 Ma (Table 1)

These two volcanic phases, but particularly phase 1, were undoubtedly the most productive of all the subaerial volcanism. The load imposed by the extrusion of considerable amounts of lava (ca. 70–80 km^3, including eroded and preserved rocks) contributed to the specific crustal mosaics subsidence and the collapse of the larger craters and their surrounding domains, with the consequent formation of *calderas* with large diameters and small depths (Figs. 3 and 4).

4.2.2. Upper complex 2 (UC2)

The UC2 construction began after the consolidation of the subaerial eruptive building during UC1 volcanism; therefore, the UC2 volcanic activity coincides with the cooling, alteration, weathering, and induration of the previously erupted lavas forming the island. The eruptive style, the number and dimension of the feeder centres and the quantity of the expelled material (ca. 10–15 km^3) during UC2 were clearly distinct from those during the UC1 volcanism.

Feeder centres	•Number: About 15 (Fig. 3) •Dimension: Changeable, but clearly subordinated to the UC1 centres. •Morphology: Mainly volcanic cones, sometimes with deep flanks; fissural structures are rare •Localization: Along the UC1 *caldera* borders and on the island periphery
Volcanic system	•A group of small-scale strato-volcanoes in association with monogenetic feeder centres
Volcanic activity	•Explosivity: 40% to 50% •Eruptive style: Oscillating between strombolian and hawaiian
Structural setting	•Fracture systems: A large number of eruptive centres are aligned or rooted on the intersection of N40°W, N20°W, N20°–30°E and NS fractures; other centres are associated to the *calderas* collapse faults
Rock record	•Sub-units E and F from UC2 (Table 2)
Time interval	•Approximately between 0.4 and 0.23 Ma (Table 1)

Because of their dominantly low viscosities, the emplacement of the UC2 lavas were largely topographically controlled. There are no strong indications (such as volcanic dikes and necks) of important intrusive activities and associated alteration phenomena during this eruptive period).

4.2.3. Upper complex 3 (UC3)

The UC3 products and structures exhibit a large contrast between the volcanic activity of this Unit and all preceding volcanism, and resulted from two distinct eruptive phases, which are recorded on the two UC3 sub-units (G-Strombolian scoria cones and H-Phreatomagmatic deposits):

Phase 1

Feeder centres	•Number: About 30 (Fig. 3) •Dimension and morphology: Small to medium scoria cones presenting usually sloppy flanks and short and smooth craters •Localization: Considerable concentration on the central area contrasting with the high dispersion on the insular periphery
Volcanic system	•Monogenetic cones
Volcanic activity	•Explosivity: 90% to 100% •Eruptive style: Strombolian, sometimes with a phreatomagmatic tendency
Structural setting	•The majority of the feeder centres are rooted on UC2 craters and vents

	•The other volcanic centres are associated to N20°E fractures
Rock record	•Sub-unit G from UC3 (Table 2).
Time interval	•Close to 0.003 Ma

Phase 2

Feeder centres	•Number: 4 centres (3 with individualized vents and 1 with 3 contiguous vents) (Fig. 3) •Dimension and morphology: Explosion craters expressing a great deepness, abrupt contours and a diameters ranging from 150 m to ca. 1 km •Localization: Inside the larger *calderas* from UC1
Volcanic activity	•Explosivity: 100% •Eruptive style: Phreatic to phreatomagmatic
Structural setting	•Fracture systems: Three northern centres are associated to N25°E and NS fractures; the southern centre is related to N40°W and N20°–25°E fractures
Rock record	•Sub-unit H from UC3 (Table 2).
Time interval	•Close to 0.0029 Ma (Table 1)

The wide dispersion and substantial lateral continuity of the UC3 phreatomagmatic deposits indicate emplacement mechanisms involving high volcanic explosivity. The pyroclastic characteristics and, particularly, the fabric of the ash deposits fabric record the progression of tephra fall and ash-cloud surges, as well as the collapse of associated ash clouds.

The very young geologic age of these events (≈0.002 Ma, in Morrisseau and Traineau, 1985) does not preclude a possible association with the present-day, but currently dormant, volcanic systems.

5. Volcanotectonic history

Taking into account (1) the volcanostratigraphy, (2) the tectonostructural evidence, particularly those related to vertical crustal movements (uplifts and subsidences) of neo- or volcano-tectonic origin and (3) the sequence of the subaerial records of marine influence, such as epiclastic marine deposits, specific erosional morphologies, intensive palagonitization and abrupt changes in the volcanic primary lithofacies (Azevedo and Ferreira, 1999), it is possible to divide the volcanotectonic evolution of Flores Island since 1 Ma ago in four major stages (Fig. 5).

5.1. Stage 1 (1.0 to 0.55 Ma)

5.1.1. Volcanic activity

It was very intensive and of the emergent type, i.e., predominantly explosive and corresponding to the transition from the proto-island to the island-formation period (Azevedo et al., 1991).

5.1.2. Crustal vertical movements

During this stage, important tectono-volcanic uplift took place, with an amplitude exceeding 100 m (the association between this uplift and the contemporaneous regression of the sea-level is evidenced by the present-day outcrops of BC formations at 400 m a.s.l.). This uplift was obviously related to the ongoing intensive volcanism of this period, though it might have been supplemented by the isostatic compensation between two adjacent lithospheric blocks (Azevedo et al., 1991; Azevedo and Ferreira, 1999), i.e., between Flores and the seamount 50 km to the west referred by Ryall et al. (1983).

5.2. Stage 2 (0.55 to 0.4 Ma)

5.2.1. Volcanic activity

Including the final eruptive phases of the UC1, the volcanic activity was characterised by the voluminous outpouring of lavas from two or three large volcanic centres. It was then followed by a long period of volcanic quiescence (0.5 to 0.4 Ma).

5.2.2. Crustal vertical movements

During this stage, the Stage-1 uplift process ceased and was followed by subsidence. The weight of the very large amounts of lava extruded during UC1 volcanism might explain the subsidence of the whole island and particularly the collapse of its central zone, with the consequent formation of volcanic *calderas*. Some stepped morphologies on NE coast of the island might be the result of subsidence along N20°E and N30–40°W up-and-down faulting throughout this stage.

5.3. Stage 3 (0.4 to 0.2 Ma)

5.3.1. Volcanic activity

This is the last major eruptive episode of this island — the UC2 volcanism.

5.3.2. Crustal movements

No important vertical tectonic movements were registered throughout this stage. However, the development of a slight volcanic doming associated with the UC2 eruptive activity cannot be dismissed.

5.4. Stage 4 (0.2 Ma until the present)

5.4.1. Volcanic activity

Following a long volcanic quiescent period (0.2 to about 0.004 Ma), a brief but very explosive volcanic episode occurred (UC3 strombolian and phreatomagmatic volcanism ≈0.003 Ma).

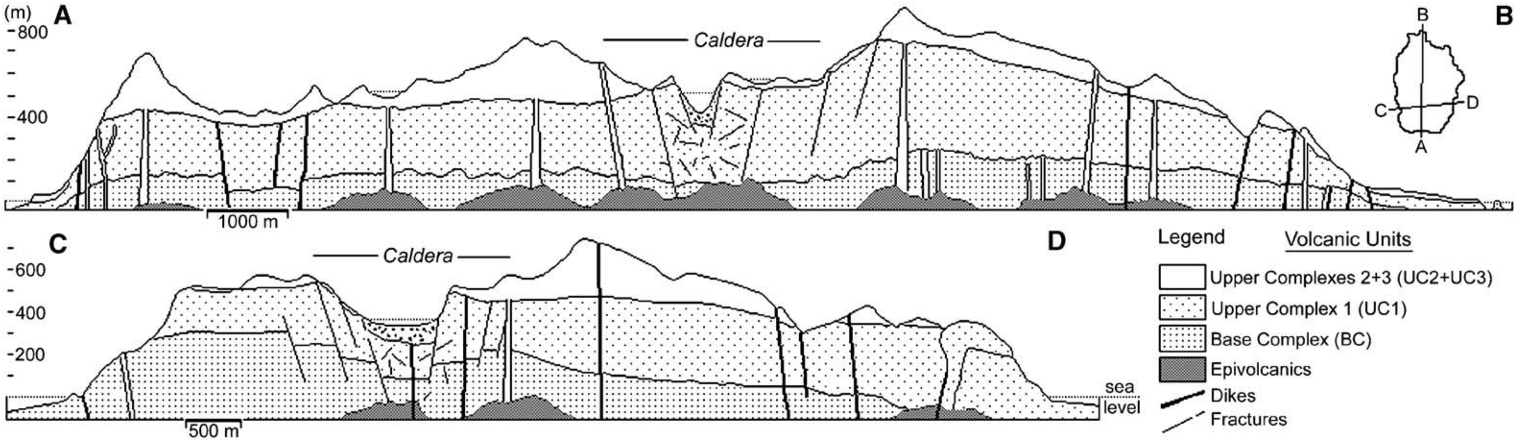

Fig. 4. Cross-sections illustrating the spatial settings of the volcanic Complexes and Units defined for Flores Island (after Azevedo, 1999).

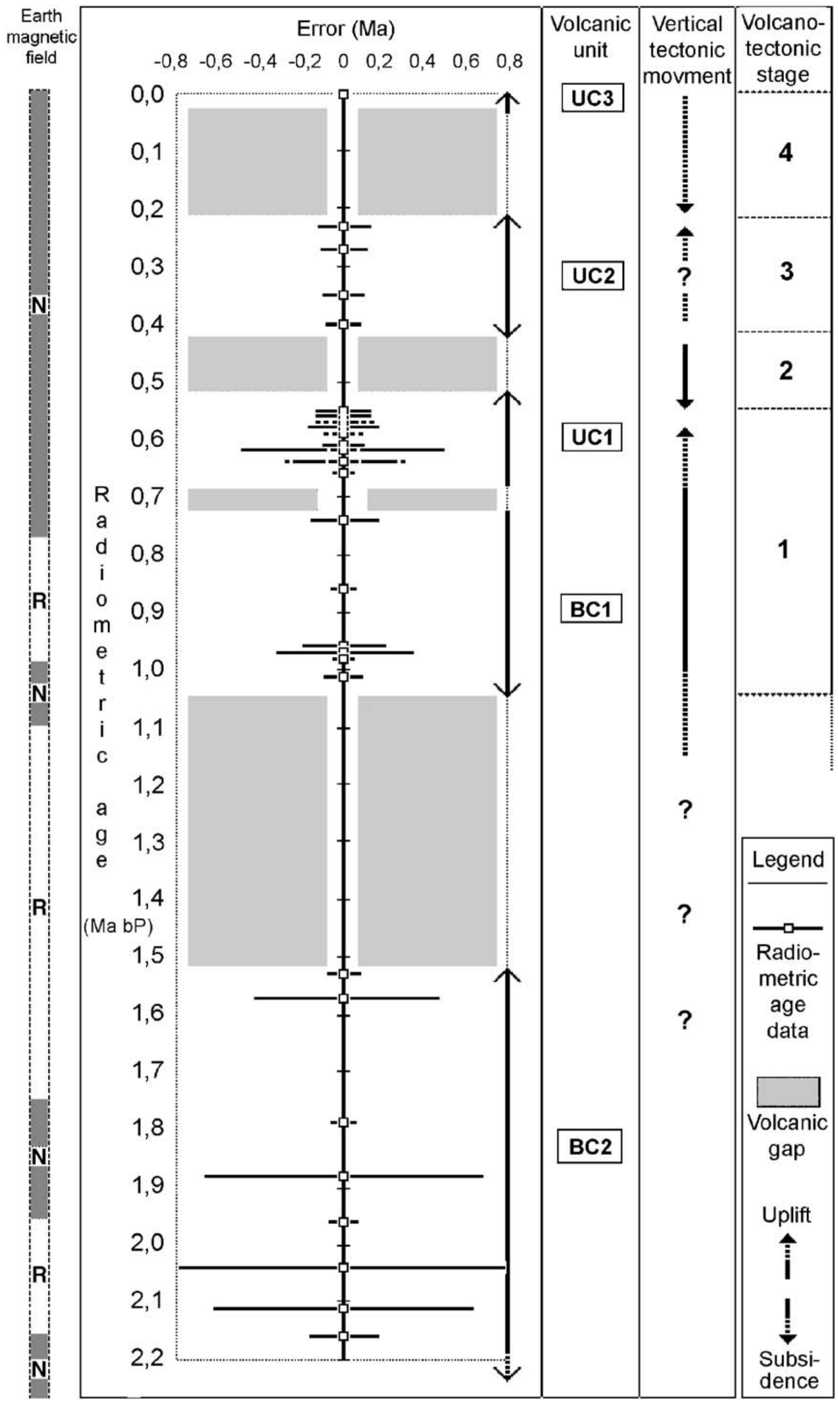

Fig. 5. Schematic representation of the radiometric ages, volcanic units, vertical tectonic movements and volcanostructural Stages defined for Flores Island (Earth's magnetic field polarity after Berggren et al., 1995).

5.4.2. Crustal movements

It is probable that the slow and gradual rate subsidence of the normally evolving oceanic crust – resulting from thermal contraction (Sclater et al., 1971) – had begun to affect the oceanic crust in Flores region, at least after the ending of the major steps of insular volcanism (UC2 and UC3). The post-insular phase might have started by then.

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